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A chronology of annual mean effective radii of stratospheric aerosols from volcanic eruptions during the twentieth century as derived from ground-based spectral extinction measurements

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Abstract. Stratospheric extinction can be derived from ground-based spectral photometric observations of the Sun and other stars (as well as from satellite and aircraft measurements available since 1979) and is found to increase after large volcanic eruptions. This increased extinction shows a characteristic wavelength dependence that gives information about the chemical composition and the effective (or area-weighted mean) radius of the particles responsible for it. Known to be tiny aerosols constituted of sulfuric acid in a water solution, the stratospheric particles at midlatitudes exhibit a remarkable uniformity of their column-averaged effective radii r_{eff} in the first few months after the eruption. Considering the seven largest aerosol-producing eruptions of the twentieth century, r_{eff} at this phase of peak aerosol abundance is ~0.3 μm in all cases. A year later, r_{eff} either has remained about the same size (almost certainly in the case of the Katmai eruption of 1912) or has increased to ~0.5 μm (definitely so for the Pinatubo eruption of 1991). The reasons for this divergence in aerosol growth are unknown.

1. Introduction

Satellite-based remote sensing of the atmosphere has revolutionized the measurement of stratospheric aerosol extinction that is produced by volcanic eruptions. Data can now be acquired almost continuously over most of the Earth's surface with both high horizontal and high vertical resolution. Nevertheless, even the best space-based data have their limitations: for volcanic aerosols the chief drawbacks of satellite observations of direct beam solar radiation at present are the coarseness of the spectral resolution available and the nonnegligible estimated errors of the retrieved absolute aerosol optical thicknesses. The satellite-based Stratospheric Aerosol Measurement II (SAM II) instrument has operated since November 1978 but measures at only one wavelength, 1.0 μm. The follow-up Stratospheric Aerosol and Gas Experiment (SAGE) system has measured at two wavelengths, 0.453 and 1.02 μm, but only from February 1979 to November 1981. Its immediate successor, SAGE II, which has functioned continuously since October 1984, measures in seven spectral bands but tabulates its retrieved aerosol extinction at four wavelengths: 0.385, 0.453, 0.525, and 1.02 μm. Uncertainties in the aerosol optical thicknesses retrieved by these various systems, however, are typically 10% at 1.02 μm and 30% at 0.385 μm for aerosols lying below 25 km [*McCormick et al.*, 1979; *Chu et al.*, 1989], while the errors can attain 100% at 0.525 μm for aerosols located above 25 km [*Wang et al.*, 1989]. On the other hand, relative errors in a single SAGE II spectral extinction curve are much smaller [*Lacis et al.*, 2000].

Stratospheric extinction data that are acquired at ground-based astronomical observatories, therefore, are still useful to have even though such data are necessarily restricted both geographically and temporally. Nevertheless, very high spectral resolution has been achieved from the ground between 0.35 and 1.6 μm, and uncertainties in the total aerosol optical thickness can be reduced to ~10% in all passbands by taking the average of a large number of observations over a restricted period of time at a superior observing site for which a long baseline of observations exists [*Stothers*, 1997]. Moreover, there is always a need for accurate ground truth measurements to validate the spacecraft observations.

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Since the 1978 launch of SAM II, two large volcanic eruptions have occurred for which the stratospheric aerosol extinction could be accurately detected from the ground by the use of either Sun photometers or stellar photoelectric photometers. The two eruptions were those of El Chichón, Mexico (17°N, March 1982), and Pinatubo, Philippines (15°N, June 1991). (The eruption of Cerro Hudson at 46°S in August 1991 produced only ~0.1 of the aerosols put up by Pinatubo [*McCormick et al.*, 1995]). The present paper makes use of published measurements of stellar extinction acquired at ground observatories to derive the mean aerosol sizes for these two eruptions. The results are then compared with available Sun photometer results derived from ground, air, and space measurements. Finally, all of these results are averaged and then added to the record of mean aerosol sizes for earlier twentieth-century eruptions [*Stothers*, 1997, 2001]. From this extensive record a number of interesting generalizations can be drawn.

However, some caveats are first necessary. To make the record homogeneous, the post-1978 data must be selected and then treated in exactly the same way as the pre-1978 data. This means that we have to exclude from the analysis all measurements based on far-infrared detectors, lidar, and in situ particle counters. All of the photometric data analyzed are ground-based atmospheric transmission measurements made simultaneously in the near ultraviolet to the near infrared; therefore, what we finally obtain are column-averaged mean aerosol sizes over just a few sites. Since time averages of the data are being utilized, however, we achieve, in effect, a limited latitudinal (and longitudinal) averaging of the data, as the volcanic aerosol cloud will drift over any given site. All sites in the present study lie near either ~30°N or ~30°S. Despite these restrictions of the data the column-averaged mean aerosol sizes at midlatitudes agree in general order with those derived by other methods, insofar as such comparisons can be made.

2. Stellar Extinction Data

In the course of making nighttime observations of stars with photoelectric photometers mounted on telescopes, astronomers routinely acquire data on atmospheric extinction which are needed to reduce the measured stellar light intensities to unobscured values lying just outside the atmosphere [*Hardie*, 1962]. The measurement procedure is similar to that used traditionally with ordinary Sun photometers. Typically, the observations of stars are made with several wideband filters that are optimally placed in the near ultraviolet to visual wavelength range. After the atmospheric extinction corrections are applied, these wholly subsidiary data are usually thrown out, although they are occasionally archived and then published in some summary form. In that case, they serve as a useful record of zenithal atmospheric extinction at the astronomical site where they were acquired. In a few cases, observations of this kind have been made with narrowband filters, including additional passbands in the red and infrared.

Published atmospheric extinction data in either Johnson's UBV or Strömgren's UVBY photometric system that cover the periods of El Chichón's and Pinatubo's eruptions are available for the following sites: La Silla, Chile [*Rufener*, 1986; *Grothues and Gochermann*, 1992; *Sterken and Manfroid*, 1992; *Burki et al.*, 1995], San Juan, Argentina [*Gil-Hutton*, 1993], and Cerro de la Virgen, Mexico [*Schuster et al.*, 1985]. Unfortunately, however, the effective wavelength range of these two standard photometric systems, $\lambda = 0.35$–0.55 μm, is too short to be decisive for determining mean aerosol sizes. Some limited observations made at Kitt Peak and at Flagstaff, Arizona, in 1981–1983 have included a red passband ($\lambda = 0.71$ μm) [*Livingston and Lockwood*, 1983; *Lockwood et al.*, 1984], but in our opinion, these measurements span too brief a time to be properly calibrated. The same problem affects the narrowband observations ($\lambda = 0.32$–0.87 μm) made during 1980 and 1984 at Cerro Tololo, Chile [*Gutiérrez-Moreno et al.*, 1982, 1986].

Winnowing of the available data leaves three large data sets that can be usefully employed. First among them is the set of narrowband observations ($\lambda = 0.33$–1.10 μm) made by *Schuster* [1982] and by *Schuster and Guichard* [1985] at San Pedro Mártir, Mexico, during the years 1973–1983. We treat 1973–1974 and 1976–1981 as undisturbed reference years. Second, there is the series of broadband UBVRI observations ($\lambda = 0.36$–0.90 μm) made by A. Landolt at Kitt Peak, Arizona, from 1960 to 1991 [*Pilachowski et al.*, 1991]. The reference period used by Landolt is 1960–1990. Third are *Kilkenny*'s [1995] $UBVR_CI_C$ observations ($\lambda = 0.36$–0.81 μm) made at Sutherland, South Africa, from 1991 on. Kilkenny used the clearest nights before August 1991 as his reference period. The first set of data will be used to study the aerosols from El Chichón, and the second and third sets will be used to study those from Pinatubo.

Extinction in the Earth's atmosphere during a volcanically disturbed period is systematically larger than that during the reference period, and the arithmetic difference between the two constitutes the volcanic aerosol turbidity perturbation. If the extinction is given in stellar magnitudes, a_λ, the optical depth follows as [*Hardie*, 1962]

$$\tau_\lambda \approx 0.921 a_\lambda. \tag{1}$$

Ultraviolet extinction measurements made from the ground are often burdened with uncertainties that show up as anomalously high values of the derived extinction with a large scatter of the individual points around the mean [*Taylor et al.*, 1977; *Stothers*, 1997]. In the present work, this particular problem affects especially *Kilkenny*'s [1995] U-band extinction, which will henceforth be ignored.

3. Method of Analysis

Some assumptions about the physical properties of the volcanic aerosols have to be made in order to invert the observed spectral extinction curve for critical information about the particle radii. In general, numerical tests already performed by many investigators [e.g., *King et al.*, 1978; *Russell et al.*, 1996] have shown that reasonable variations in these physical assumptions do not lead to significant differences in the results of the analysis for the effective particle radius. Nevertheless, the adopted assumptions are explicitly stated here: a spherical particle shape, a lognormal distribution of particle radii, a composition of 75% H_2SO_4 and 25% H_2O, and a complex refractive index given by $m = 1.43 - 0i$. Ash particles can be ignored because the bigger ones fall out of the atmosphere in a few days or weeks following the eruption and the rest show little discernible optical effect after that [*Russell et al.*, 1996; *Schneider et al.*, 1999]. The lognormal distribution, which has been recommended for general use in recent surveys of volcanic aerosols in the stratosphere [*Lenoble and Brogniez*, 1984; *Russell et al.*, 1993], is given by

$$n(r) = N(2\pi)^{-1/2}\,(r \ln \sigma_1)^{-1} \exp\,[-\ln^2(r/r_1)/(2 \ln^2 \sigma_1)], \tag{2}$$

where $n(r)$ is the concentration of particles with radii between r and $r + dr$, N is the total number of particles, r_1 is the (somewhat misleadingly named) modal radius, and σ_1 is the geometric standard deviation of the radii.

Mie-scattering theory will be used to calculate the scattering efficiency factor Q. This quantity is needed in order to construct theoretical aerosol extinction curves, of which a large set is generated here. These curves are then fitted to the observed curve by applying the standard method of least squares, and the theoretical curve that produces the smallest residual is deemed the best fit solution [*Stothers*, 1997]. (Equation (A7) of *Stothers* [1997] for the theoretical τ_λ contains a misprint: $(2\pi/\lambda)^2$ should be $(2\pi/\lambda)^i$.) The theoretical curves are uniquely defined by the two parameters r_1 and σ_1. In practice, if the observed curve is not exceptionally well measured, the residuals may turn out to be approximately constant over a wide range of σ_1 values extending from unity (the minimum possible value) up to some σ_1^*. Over this range the best fit values of r_1 are then found to decrease more or less monotonically with increasing σ_1 in such a way that the effective (area-weighted mean) radius $r_{\rm eff}$ remains nearly constant [*Lacis et al.*, 1992]. The latter quantity is defined very generally by

$$r_{\rm eff} = \int n(r)\, r^3\, dr \Big/ \int n(r)\, r^2\, dr. \tag{3}$$

For a unimodal lognormal distribution, $r_{\rm eff}$ is related to r_1 and σ_1 by

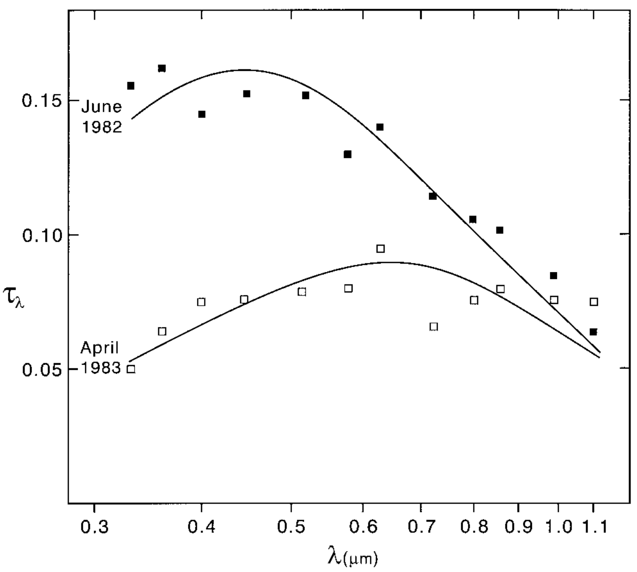

Figure 1. Theoretical spectral extinction curves fitted to the stratospheric optical depth perturbations detected over San Pedro Mártir, Mexico, in June 1982 and in April 1983.

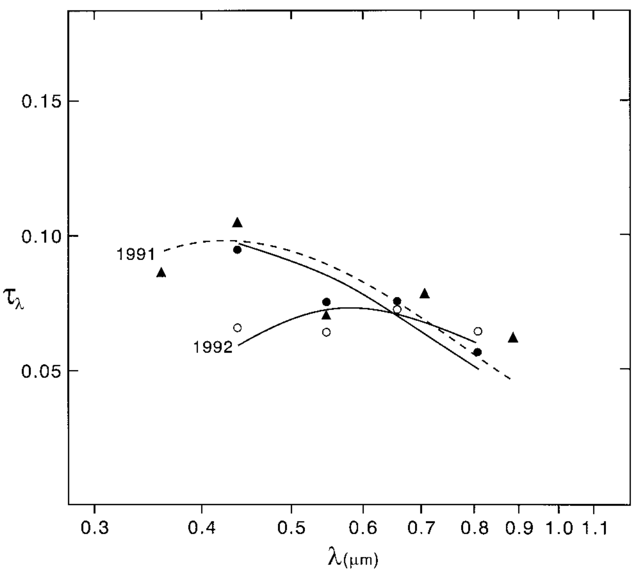

Figure 2. Theoretical spectral extinction curves fitted to the stratospheric optical depth perturbations detected over Kitt Peak, Arizona, in late 1991 (dashed curve and triangles) and over Sutherland, South Africa, in late 1991 (solid curve and solid circles) and in 1992 (solid curve and open circles).

$$r_{\rm eff} = r_1 \exp\,[(5/2)\,\ln^2 \sigma_1]. \tag{4}$$

Results will be presented here both for the monodisperse case $\sigma_1 = 1$ and for the special case $\sigma_1 = 1.3$ because the latter value is found to be typical of volcanic aerosols in the stratosphere [*Wang et al.*, 1989; *Deshler et al.*, 1993; *Stenchikov et al.*, 1998]. Sometimes volcanic aerosol distributions are observed to be bimodal, but even in such cases a unimodal lognormal assumption usually gives a sufficiently accurate approximation if the modes overlap [*Russell et al.*, 1996]. Since $\sigma_1^* > 1.3$, the dependence of $r_{\rm eff}$ on σ_1 turns out to be very slight for the present data. Denoting by r_M the true mode of the particle radius distribution, we have for the lognormal distribution

$$r_M = r_1 \exp\,(-\ln^2\sigma_1). \tag{5}$$

4. Effective Particle Sizes After El Chichón and Pinatubo

Stratospheric optical depth perturbations that were measured at San Pedro Mártir (31°N) in June 1982 and in April 1983 are plotted against wavelength in Figure 1. Best fit theoretical curves are also shown. Within the scatter of the observational points the theoretical fits seem to be satisfactory and imply $r_{\rm eff} = 0.36$ μm 3 months after the El Chichón eruption and $r_{\rm eff} = 0.51$ μm ~1 year later.

Sun photometer measurements that were made either from the ground or from high-flying aircraft have been analyzed by several authors. Using the nonparametric inversion method of *King et al.* [1978, 1984], *Dutton et al.* [1994] found $r_{\rm eff} = 0.40$ μm at Mauna Loa, Hawaii (20°N), in September 1982. From aircraft observations at 35°S–44°N during October and November of the same year, *Spinhirne and King* [1985] derived an average $r_{\rm eff} = 0.36$ μm. These values agree well with our June 1982 value for San Pedro Mártir derived from *Schuster*'s [1982] data.

In March–May 1983, $r_{\rm eff}$ still hovered near 0.36 μm at the following locations: Mauna Loa (20°N) [*Dutton et al.*, 1994], the Pacific Ocean (55°S–56°N, from aircraft) [*Spinhirne and King*, 1985], and Bedford, Massachusetts (42°N) (F. Volz, personal communication, 1996). The San Pedro Mártir value for April 1983, however, is much larger, 0.51 μm.

This puzzling dichotomy for $r_{\rm eff}$ can be seen also in published values of the mode r_M of the particle radius distribution in cases where the values of $r_{\rm eff}$ have not been published. For the period January–July 1983, r_M turns out to have been 0.3 μm at Hamilton, Ontario (43°N) [*Davies et al.*, 1988], 0.35 μm at Tsukuba, Japan (36°N) [*Asano et al.*, 1985], 0.4–0.5 μm at Monte Cimone, Italy (44°N) [*Levizzani and Prodi*, 1988], 0.5 μm at Sendai, Japan (38°N) [*Shiobara et al.*, 1991], and 0.6 μm at Richland, Washington (46°N) [*Michalsky et al.*, 1984, 1990].

The principal cause of the apparent disagreements for 1983 is probably large-scale patchiness of the aerosol cover. Although each derived value of the effective (or mode) radius is based on several days' data and on a broad range of measured wavelengths (~0.4 to ~1 μm), monthly spectral extinction differences can still be pronounced at different latitudes, even many months after the eruption, as is shown by *Dutton and DeLuisi*'s [1983] aircraft surveys made in December 1982. It must also be kept in mind that the inversion technique of *King et al.* [1978] is a nonparametric method that is sensitive to the specified total range of the particle radii and to the specified number of iterations for a solution.

Turning now to the eruption of Pinatubo, the stratospheric optical depth perturbations over Kitt Peak (32°N) and over Sutherland (32°S) in late 1991 and again over Sutherland in 1992 are plotted in Figure 2. The best theoretical fits to the 1991 data occur if $r_{\rm eff} = 0.32$ μm at Kitt Peak and if $r_{\rm eff} = 0.31$ μm at Sutherland. These stellar extinction results show excellent interhemispheric agreement. An independent analysis of the Kitt Peak data by *Kerola and Timmermann* [1992] produced a very similar result, $r_{\rm eff} = 0.35$ μm. For the following year the Sutherland data yield $r_{\rm eff} = 0.52$ μm.

These values can be compared with previously published Sun photometer results, derived mostly by the method of *King*

Table 1. Effective Radii of Volcanic Aerosols Derived From Ground-Based Measurements

Volcano	Latitude	Longitude	Eruption Date	Measurement Dates	Measurement Latitude	r_{eff},[a] μm	r_{eff},[b] μm
Santa Maria	15°N	92°W	Oct. 1902	1903–1904	39°N	0.31	0.34
Ksudach	52°N	158°E	March 1907	May 1907	39°N	0.24	0.26
				June–Oct. 1908	34°N	0.46	0.46
Katmai	58°N	155°W	June 1912	July–Aug. 1912	34°N	0.33	0.34
				Aug.–Oct. 1913	34°N	0.31	0.34
				June–Oct. 1914	34°N	0.36[c]	0.37[c]
Agung	8°S	116°E	March 1963	May–Dec. 1963	29°S	0.35	0.37
				May 1964	29°S	0.30	0.32
Fuego	14°N	91°W	Oct. 1974	1975	31°N	0.39	0.41
El Chichón	17°N	93°W	March 1982	June 1982	31°N	0.36	0.37
				April 1983	31°N	0.51[d]	0.48[d]
Pinatubo	15°N	120°E	June 1991	Oct. 1991	32°N	0.32	0.34
				Aug.–Dec. 1991	32°S	0.31	0.33
				1992	32°S	0.52	0.46

[a]Assuming $\sigma_1 = 1.3$.
[b]Assuming monodisperse particles ($\sigma_1 = 1$).
[c]Uncertain value.
[d]Not necessarily typical; some other localities show $r_{\mathrm{eff}} \approx 0.36$ μm.

et al. [1978] or, as in the case of the results of *Saxena et al*. [1995] and of *Hansen et al*. [1996], by two independent methods. In mid to late 1991, r_{eff} was 0.18–0.35 μm at 10°–15°N (from aircraft) [*Valero and Pilewskie*, 1992], ~0.31 μm at two Antarctic stations (67°–71°S) [*Herber et al*., 1996], ~0.35 μm at Trivandrum, India (9°N) [*Moorthy et al*., 1996], ~0.4 μm at 50°–70°S (SAGE II) [*Saxena et al*., 1995; *Hansen et al*., 1996], 0.55 μm at Mauna Loa, Hawaii (20°N) [*Russell et al*., 1993; *Dutton et al*., 1994], and ~0.58 μm at Hyderabad, India (17°N) [*Ramachandran et al*., 1994]. In their analysis of the Mauna Loa data, *Russell et al*. [1993] found that the nonparametric method of King et al. produces larger effective radii than a simple lognormal assumption does and that adopting the latter assumption yields $r_{\mathrm{eff}} = 0.44$ μm. Also, there can be little size discrimination in any method for $r_{\mathrm{eff}} \gtrsim 0.5$ μm [*Lacis et al*., 2000]. All in all, the discrepancies occurring for the Mauna Loa and Hyderabad data may be more apparent than real, and the overall evidence strongly favors a value of r_{eff} near ~0.35 μm.

In the 1992 case the published Sun photometer results scatter in a similar way, r_{eff} being ~0.43 μm at two Antarctic stations (67°–71°S) [*Herber et al*., 1996], ~0.5 μm at 30°–60°N (SAGE II) [*Anderson and Saxena*, 1996; *Hansen et al*., 1996], 0.57 μm [*Dutton et al*., 1994] or 0.86 μm [*Russell et al*., 1993] at Mauna Loa, Hawaii (20°N), ~0.6 μm at 20°–90°N (from aircraft) [*Pueschel et al*., 1994], ~0.60 μm at Trivandrum, India (9°N) [*Moorthy et al*., 1996], and ~0.65 μm at Tsukuba, Japan (36°N) [*Asano et al*., 1993]. The anomalously high value of 0.86 μm derived by *Russell et al*. [1993] for the Mauna Loa data is reduced to 0.57 μm under a lognormal assumption; this much lower value coincides well not only with the analysis of the Mauna Loa data by *Dutton et al*. [1994] but also with the value of ~0.50 μm at 0°–80°N (from aircraft) that was independently derived under a lognormal assumption by *Russell et al*. [1996]. The weight of evidence therefore points to an average value near ~0.55 μm for 1992.

In situ particle counters have tended to find slightly smaller mean particle radii than those derived by spectral extinction methods. In late 1982, a half year after El Chichón erupted, the stratospheric particle counts peaked at $r_M = 0.2$–0.3 μm [*Hofmann and Rosen*, 1983, 1984; *Knollenberg and Huffman*, 1983; *Oberbeck et al*., 1983; *Snetsinger et al*., 1987], just as they did a year after the 1963 Agung eruption [*Mossop*, 1964]. For $\sigma_1 = 1.3$ this peak corresponds to $r_{\mathrm{eff}} \approx 0.30$ μm, using (4) and (5). Following the Pinatubo eruption, the particle counts during late 1991 and all through 1992 [*Deshler et al*., 1992, 1993; *Wilson et al*., 1993; *Pueschel et al*., 1994; *Goodman et al*., 1994] again yielded somewhat smaller than expected r_M values. Various explanations of this discrepancy have been suggested by *Russell et al*. [1996], chief among them being (1) the possibility that the particle sampling has missed most of the very few, very large, optically effective particles and (2) the complementary tendency of spectral extinction methods to underestimate the number of very small particles. Furthermore, in situ observations have sampled only the lowermost stratosphere, in which the local mean particle sizes might be smaller than the total column-averaged ones [*Grainger et al*., 1995].

Another method of inferring the dominant particle radius is through measurement of the angular size of Bishop's ring around the Sun, a colored halo phenomenon that has often been seen after large volcanic eruptions. The problem is that the ring's angular size, and hence the inferred dominant particle radius, 0.6–0.8 μm, appears to be nearly invariant in time and space [*Stothers*, 1996]. Owing to such interpretational difficulties, we do not consider this method further.

5. Twentieth-Century Chronology of Effective Particle Sizes

Effective particle radii derived from ground-based spectral extinction measurements using the reference year method and assuming a unimodal lognormal distribution of particle radii are listed in Table 1 for the largest aerosol-producing twentieth-century volcanic eruptions. Data sources are the following: spectrobolometric observations of the Sun by members of the Astrophysical Observatory of the Smithsonian Institution for Santa Maria (1902), Ksudach (1907), and Katmai (1912) [*Stothers*, 1996]; multicolor photoelectric photometry of stars by *Irvine and Peterson* [1970] for Agung (1963) and by *Schuster* [1982] for Fuego (1974), and as very recently analyzed by

Stothers [2001]; and multicolor stellar photometry for El Chichón (1982) and for Pinatubo (1991), as analyzed in this paper.

When the visual optical depth perturbation in the stratosphere drops below ~0.03, these ground-based methods are rather inaccurate. Therefore, during the later years following a large eruption, either spacecraft or aircraft observations are desirable. These are, in fact, available after both El Chichón (during 1984) [*Wang et al*., 1989] and Pinatubo (during 1993 and 1994) [*Stone et al*., 1993; *Anderson and Saxena*, 1996; *Hansen et al*., 1996; *Russell et al*., 1996].

What can we now conclude about r_{eff} in the early posteruption years? As soon as the original aerosols have formed in the stratosphere, their growth proceeds by coagulation and mutual collisions until a kind of quasi-steady state emerges a few weeks after the eruption. After that period, although the rate of growth slows, the size distribution begins to be altered by horizontal dispersion and by gravitational and convective sedimentation of the larger particles. There is little a priori reason therefore to suspect that different volcanic eruptions, with their very dissimilar injections of SO_2 into the stratosphere in regard to quantity, height, and latitude, would lead to virtually the same column-averaged effective particle radius at midlatitudes after the first few weeks. Yet the tabulated eruptions do so: r_{eff} is clearly ~0.3 μm in all cases. Even the small eruption of Nevado del Ruiz, Colombia (5°N, November 1985) led to the same r_{eff} [*Lacis et al*., 2000].

A year after the eruption, some of the matured aerosol veils have shown little evolution of their column-averaged effective particle radii, while others, like the veils from Ksudach and Pinatubo, have apparently experienced a large increase of r_{eff} to ~0.5 μm. The explanation for this varied behavior is uncertain but may be simply the fact that marked patchiness of the veils sometimes occurs. On the other hand, broad swaths of latitude were surveyed from aircraft a year after El Chichón and from spacecraft a year after Pinatubo, and broadly speaking, r_{eff} at midlatitudes in those studies turned out to average ~0.36 μm and ~0.5 μm, respectively. Also, it seems difficult to attribute these large differences in r_{eff} solely to variations in analytical methodology since for most eruptions the same analysis technique was used, and, in fact, even a brief glance at the slope of an observed spectral extinction curve can usually provide a fairly good estimate of r_{eff} [*Lacis and Mishchenko*, 1995]. Finally, the magnitude of a volcanic eruption does not appear to be a reliable predictor of r_{eff}. For although Pinatubo, Santa Maria, and Agung generated comparable amounts of aerosol, i.e., 20–30 Tg, Ksudach and Fuego put up only 2–5 Tg, and Katmai and El Chichón put up 10–15 Tg. We tentatively conclude that the similarities and differences in r_{eff} are real, even if presently not understood.

The whole question about establishing accurately the value of r_{eff} is important because just two quantities, r_{eff} and the visual optical depth perturbation, provide essentially all of the empirical information that is required to compute the atmospheric radiative forcing by a stratospheric aerosol cloud [*Lacis et al*., 1992; *Lacis and Mishchenko*, 1995]. Refined values of the visual optical depth perturbation are now available for the years 1881–1978 [*Stothers*, 1996, 2001] and 1979 to the present [*Sato et al*., 1993; M. Sato and J. Hansen, personal communication, 2000], and these optical depths are consistent with the present chronology of r_{eff} because they are based on the same, or very similar, data. There are several uses of such data in theoretical climate studies. As tracers, volcanic aerosols track the dynamics and mixing of the stratosphere. They also affect the chemistry of the stratosphere, e.g., the ozone abundance. In addition, they have indirect effects on temperature and precipitation near the ground. As a result, long-term global climate modeling that covers periods of intense volcanism must depend sensitively on a knowledge of the various ways in which volcanic aerosols can affect the whole atmosphere.

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